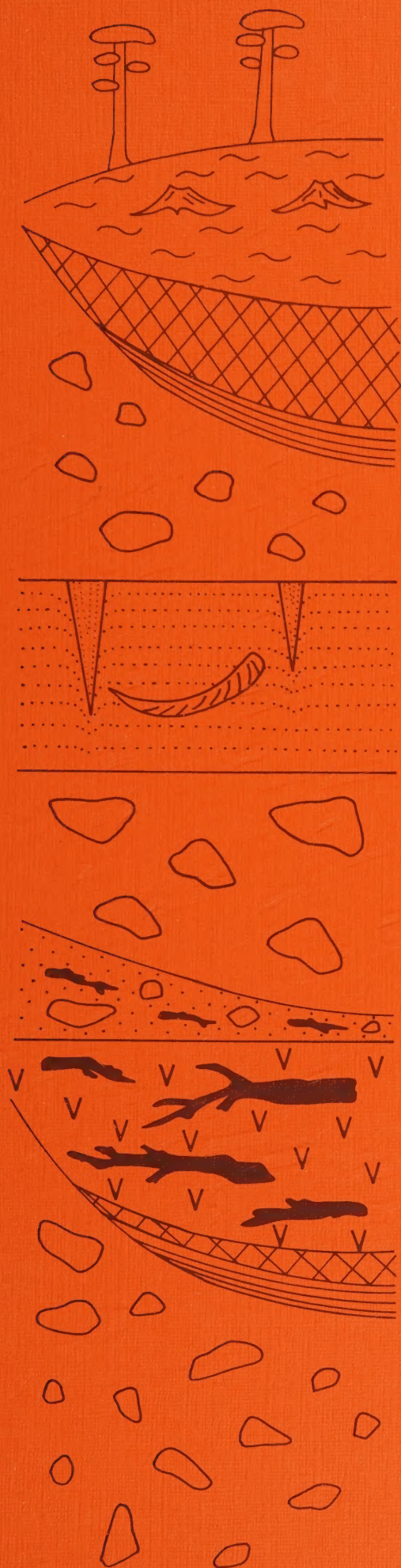




UNIVERSITY OF LUND
DEPARTMENT OF
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STUDIES OF THE QUATERNARY IN NORTHEAST GREENLAND

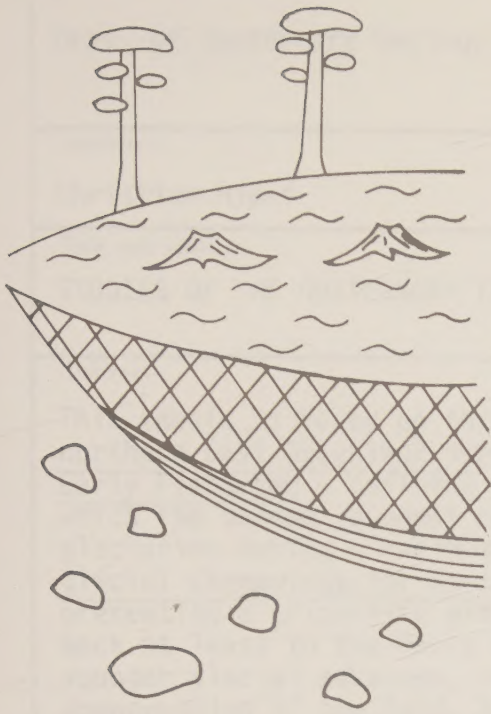
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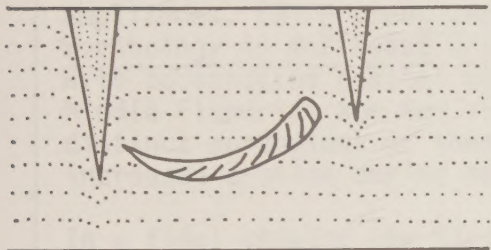
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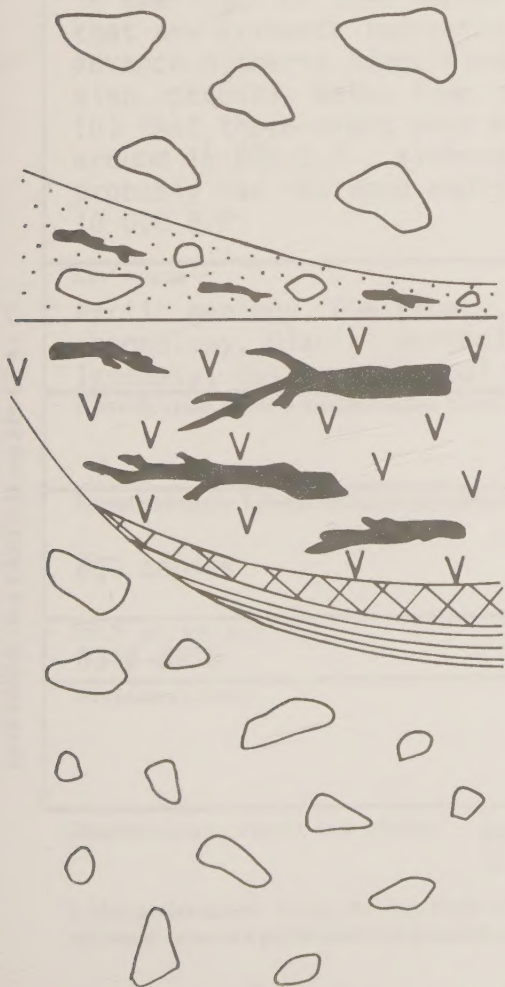
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Abstract:

This thesis is based on three papers: (1) Glaciation in northern East Greenland during the Late Weichselian and Early Flandrian - defining three glaciation zones of which the second youngest reflects the rather limited glaciation during a maximum around 10 000 B.P. (2) A glacial chronology for northern East Greenland - presenting a tripartite glacial chronology which goes back at least to the Early Weichselian and where the younger glacial advances, and the corresponding isostatic downpressing of the land, had successively smaller amplitudes. (3) Present and Middle Flandrian coastal morphology in Northeast Greenland - describing two zones with accentuated coastal morphology, an elevated one of Middle Flandrian age and the present coastline. The distinct morphology of these zones is attributed to increased erosion during marine transgressions.

In a final chapter the first two papers are discussed in the light of some recent studies, and it is noted that new evidence indicates (a) that the oldest glacial advance hitherto identified, that of the greatest extension, probably dates from the Saale glaciation, and (b) that there might have been a glacial advance also around 18 000 B.P., although in most areas it then probably had the same amplitude as the one around 10 000 B.P.

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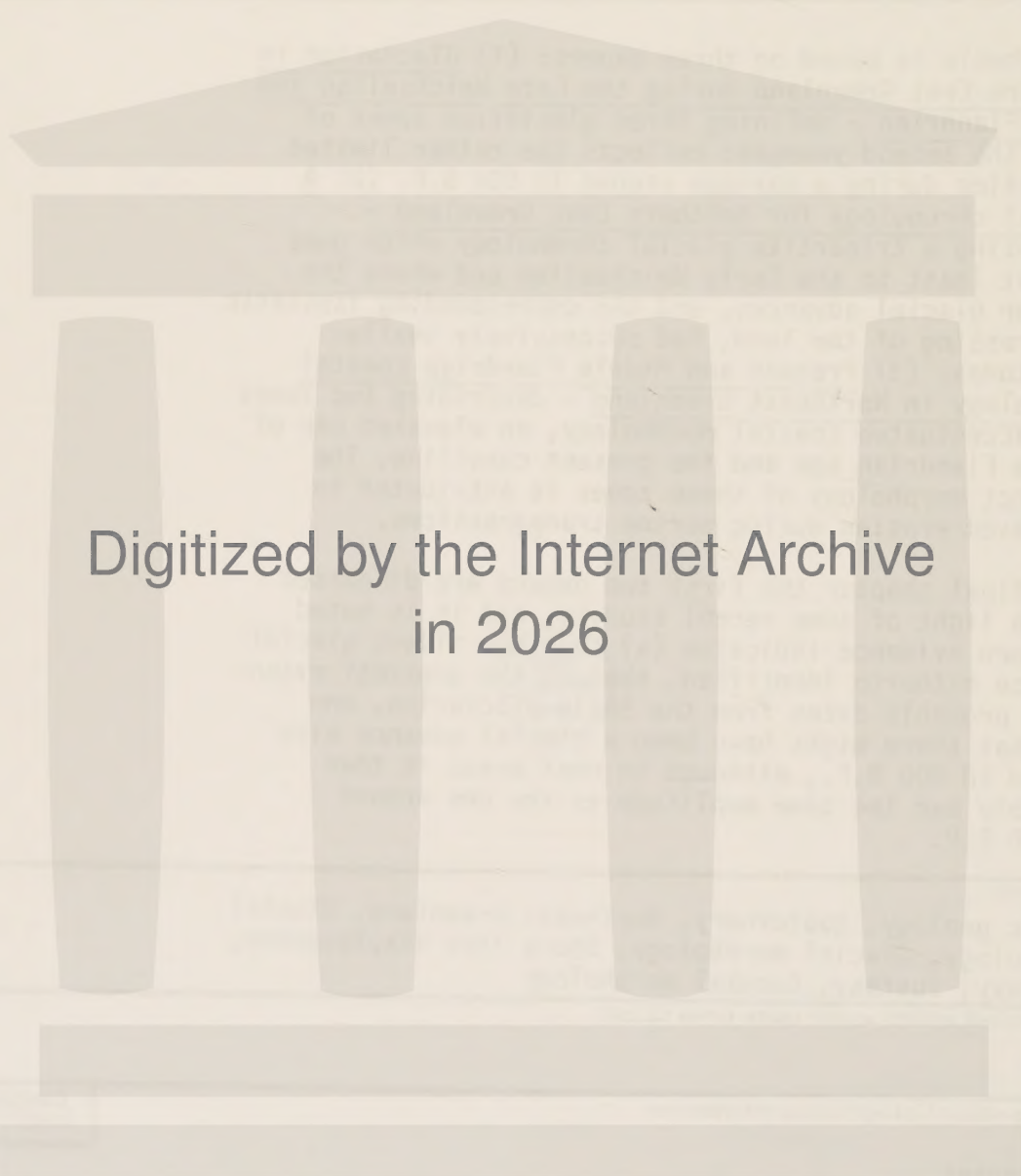
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1. INTRODUCTION

This thesis is based on the following three papers. They are summarized in chapter 3 where, however, only a limited number of the most significant references mentioned in the original papers are given, mainly those directly related to arctic studies.

- (1) Hjort, Christian 1979: Glaciation in northern East Greenland during the Late Weichselian and Early Flandrian. *Boreas*, Vol. 8, pp. 281-296.
- (2) Hjort, Christian 1981: A glacial chronology for northern East Greenland. *Boreas*, Vol. 10, pp. 257-272.

- (3) Hjort, Christian 1981: Present and Middle Flandrian coastal morphology in Northeast Greenland. *Norsk Geografisk Tidsskrift*, Vol. 35, in press.

These are some results of my field work in Northeast Greenland, which started in 1970. According to the original research plan that work was intended to be a conventional study of the deglaciation chronology within the area between Kong Oscars Fjord and Kejser Franz Josefs Fjord (Fig. 1), based on the then common assumption that during the later part of the Weichselian glaciers had covered almost everything. But it soon turned out that the Quaternary geology of Northeast Greenland was more complicated than that, and this applied to my own study area as well as to the Scoresby Sund region in the south, where my Danish colleague Svend Funder worked. A first preliminary paper, showing the way towards a chronological picture with a limited late Weichselian glaciation and a very early maximum glaciation, was published in 1973 (Funder & Hjort 1973). Since then we have worked our way northwards, and in the summer of 1979 even sampled parts of Oodaaq Ø north of Peary Land, at approximately 83°40'N, the world's most northerly land. The new material from North Greenland and the northernmost parts of the east coast, including lots of very old shell-bearing sediments and even a more or less intact interglacial sequence 100 m thick (Funder & Hjort 1980, Hjort 1981), seems even more promising than the material from those parts of the east coast dealt with here. This thesis and the papers summarized here, which include material from 1970-1976, should thus only be seen as stepping stones towards a greater understanding of the Quaternary of the whole northeastern part of Greenland.

2. DATING METHODS AND TERMINOLOGY

Shells from almost all the shell-bearing deposits discussed in the three papers have been ^{14}C dated, by Sören Håkansson at the Radiocarbon Dating Laboratory in Lund, Sweden. Some older samples were ^{14}C dated a second time by Reidar Nydal at the Radiological Dating Laboratory in Trondheim, Norway. Three older samples from Hochstetter Forland and Shannon were also Th/U dated in Trondheim. A total of 9 samples have been amino acid analysed, as to the ratio of isoleucine epimerization (Allo:Iso ratio), by Gifford H. Miller at the Institute of Arctic and Alpine Research in

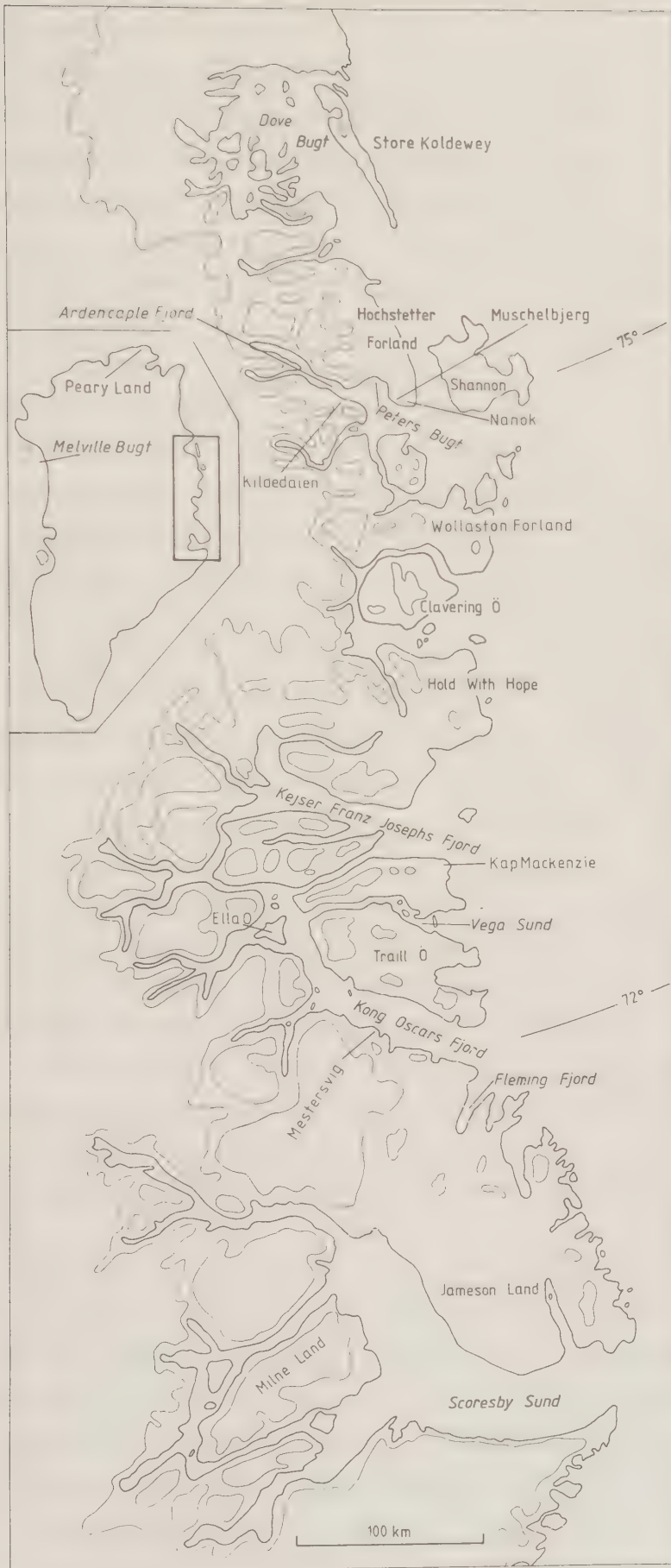


Fig. 1. Northeast Greenland, with the names of localities mentioned in the text. A broad outline of the topography is given by the coastline and the 1000 m contour line. Most of the areas above 1000 m lie within the Caledonides, the extension of which is thus illustrated.

Boulder, Colorado, USA. The aim of the ^{14}C and Th/U datings has of course been to get ages as absolute as possible, although in the case of several ^{14}C datings they are clearly minimum ages. The aim of the amino acid analyses has been to check on the stratigraphy, that is to see whether the samples which are or seem to be the oldest ones stratigraphically really are oldest, and also to group the sediments into approximate age classes. The method seems to work well in this respect, but the results may indicate the presence of a few older shells in some of these shell-bearing marine sediments.

The terminology used is mainly that proposed by Mangerud et al. (1974), but with the addition of a tentative lower limit for the Weichselian (Wisconsin) at 120 000 B.P., in accordance with the usage of Andrews & Barry (1978) and Boulton (1979) for Baffin Island and Spitsbergen. I have, for the time being, disregarded the recommendations of e.g. Mangerud & Berglund (1978) and the INQUA Subcommission for European Quaternary Stratigraphy (Chaline et al. 1980) to use 25 000 B.P. instead of 13 000 B.P. as the lower limit of the Late Weichselian. This conservatism seemed wise in the case of Northeast Greenland when no signs of any glacial advance around 22 000 - 18 000 B.P. had been found, but, as will be evident from chapter 4, this may change.

I have also retained the term Flandrian, as a convenient and widely used but perhaps slightly informal stage name for the present interglacial. My arguments for doing so are mainly those of Hyvärinen (1978) and West (1979). It may be added that in fact there seems to be no good alternative. The commonly used term "Holocene" gives too high a rank to what is obviously no more than an ordinary interglacial within the Pleistocene epoch. The fact that the deposits laid down during the later part of this interglacial have to a large extent been characterized by the effects of an overpopulation of the Earth by our own species should not lead us to exaggerate its geological importance. Should man be used at all as a time marker, it would seem more sophisticated to follow the Russian approach and join the Pleistocene and Pliocene into the Anthropogene (e.g. Gerasimov 1979). Especially as we now know that the first signs of Cenozoic glaciation in Antarctica probably go back to the Oligocene (Kvasov & Verbitsky 1981), and glaciation in the Northern Hemisphere at least to the early Pliocene (e.g. Margolis & Herman 1980).

3. SUMMARIES OF THE PAPERS

3.1 Glaciation in northern East Greenland during the Late Weichselian and Early Flandrian

This paper is concerned with the extent of glaciation in northern East Greenland during the Late Weichselian and Early Flandrian. It is mainly based on my own work from the Kong Oscars Fjord area and northwards, but some data from earlier authors are included, and all data from the Scoresby Sund region and some from the area between Kejser Franz Josefs Fjord and Wollaston Forland derive from the work of Svend Funder (e.g. Funder 1978).

Three different glaciation zones are defined: Zone A, the youngest one, is the very fresh and largely unvegetated zone immediately in front of, and above, the present glaciers and ice caps. It has been exposed during the withdrawal from the maximum positions which these glaciers held in the early part of this century. Zone B is the second youngest and much more extensive than Zone A. In this zone a profusion of glacial deposits and traces of glacial erosion bear witness to a time when almost all larger valleys and fjord basins were inundated by glaciers, sometimes reaching the outer coast. Within this zone, since long completely colonized by vegetation, the glacial landforms are more subdued than in Zone A, but still largely intact. Zone C is the upper- and outermost and oldest zone. Here the glacial landforms have often been more or less worn down by erosion. The rocks are extensively weathered, tors occur and felsenmeers may cover large areas. Striae and other traces of glacial erosion sometimes occur, but the best proof that also these areas were at one time covered by active ice is usually the occurrence of erratic blocks. These are found up to high altitudes, even in the coastal mountains, showing that an ice-sheet reaching onto the shelf once covered the whole area.

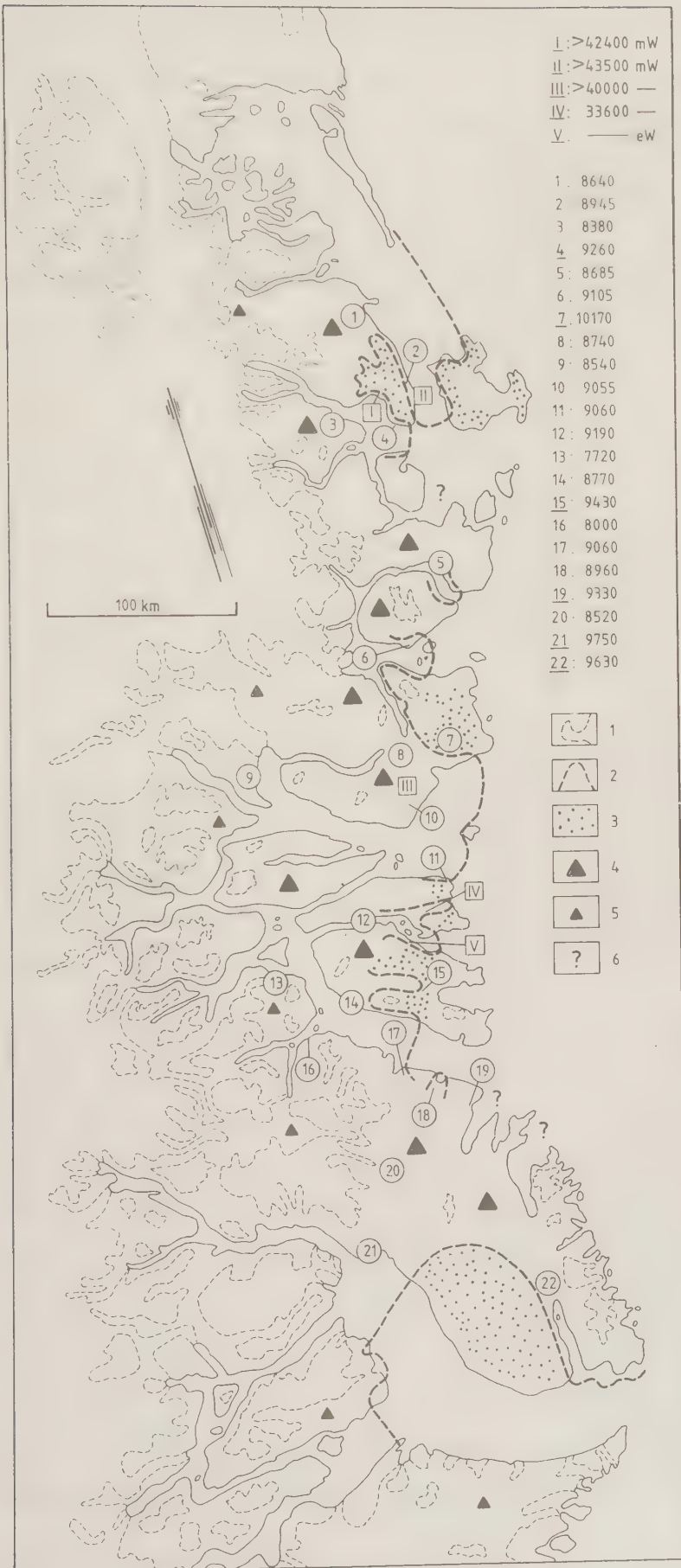
The paper is then mainly concerned with the outer and upper boundary of Zone B, called the B-line. This line is often morphologically delimited by distinct terminal or lateral moraines, kame terraces and lateral drainage channels, or by terminal or lateral deltas. Where the morphology is less distinct the B-line may be determined by an increase upwards or outwards in the degree of weathering, and/or by a distinct decrease in the number of glacial erratics. The extent of glaciation when the ice stood at the

B-line is here shown as Fig. 2 (= Fig. 2 in Paper 1). As can be seen, large ice-free lowland areas as well as a sizeable number of nunataks existed at that time.

The formation of the B-line during an independent glacial advance is concluded from the following facts: (1) The striking difference in morphological maturity and weathering, and in the frequency of glacial erratics, between areas on different sides of this boundary. (2) The well developed and well preserved lateral moraines which often mark the upper limit of Zone B. (3) The existence of Early as well as Middle Weichselian shell fragments in the deposits laid down by the glaciers reaching the B-line. (4) Studies of glacial striae indicate the existence of a period with only local glaciation, preceding that glacial period when the ice stood at the B-line. (5) The B-line is contemporaneous with, or predates, a distinct marine limit above which only scattered traces of distinctly older marine deposits occur. The altitudes within the region of this younger marine limit, and its relation to older marine deposits, is here shown as Fig. 3 (= Fig. 10 in Paper 1).

The B-line has been dated to around 10 300 B.P. on the Hold With Hope peninsula north of Kejser Franz Josephs Fjord, and at the Milne Land moraines in Scoresby Sund. From Fleming Fjord and central Traill Ø come datings giving approximate ages of the local marine limit of 9330 B.P. and 9430 B.P., respectively. These may however slightly post-date the glacial maximum.

Fig. 2. The extent of present glaciation, and the situation during the Late-Weichselian-Early Flandrian glaciation maximum. (1) Present glaciation. (2) The maximum positions of glacier fronts in the fiords, sounds and larger valleys around 10,000 B.P. (3) Lowland areas which have been unglaciated at least since the Middle Weichselian, and lower plateau areas which have been at least partly unglaciated. (4) Zone with abundant occurrences of nunataks during the glaciation maximum around 10 000 B.P. (5) Zone with less abundant occurrences of nunataks at that time. (6) Position of glacier fronts unknown. Also shown are a number of dated samples. Nos. I-V (marine shells) indicate the absolute maximum ages for the onset of glacial advance and show some minimum distances covered by it. Both ^{14}C ages (uncorrected) and Amino Acid ages (mW - middle Weichselian, eW - early Weichselian) are given. Samples 1-21 are ^{14}C dated marine shells corrected for apparent age; No. 22 is an uncorrected lake sediment sample. Samples 7 and 21, perhaps also 15 and 19, are relevant for dating the glacial maximum. The other samples in that series date the oldest known shell-bearing deposits in different areas (No. 22 the approximate onset of sedimentation), thus giving minimum ages for the deglaciation.



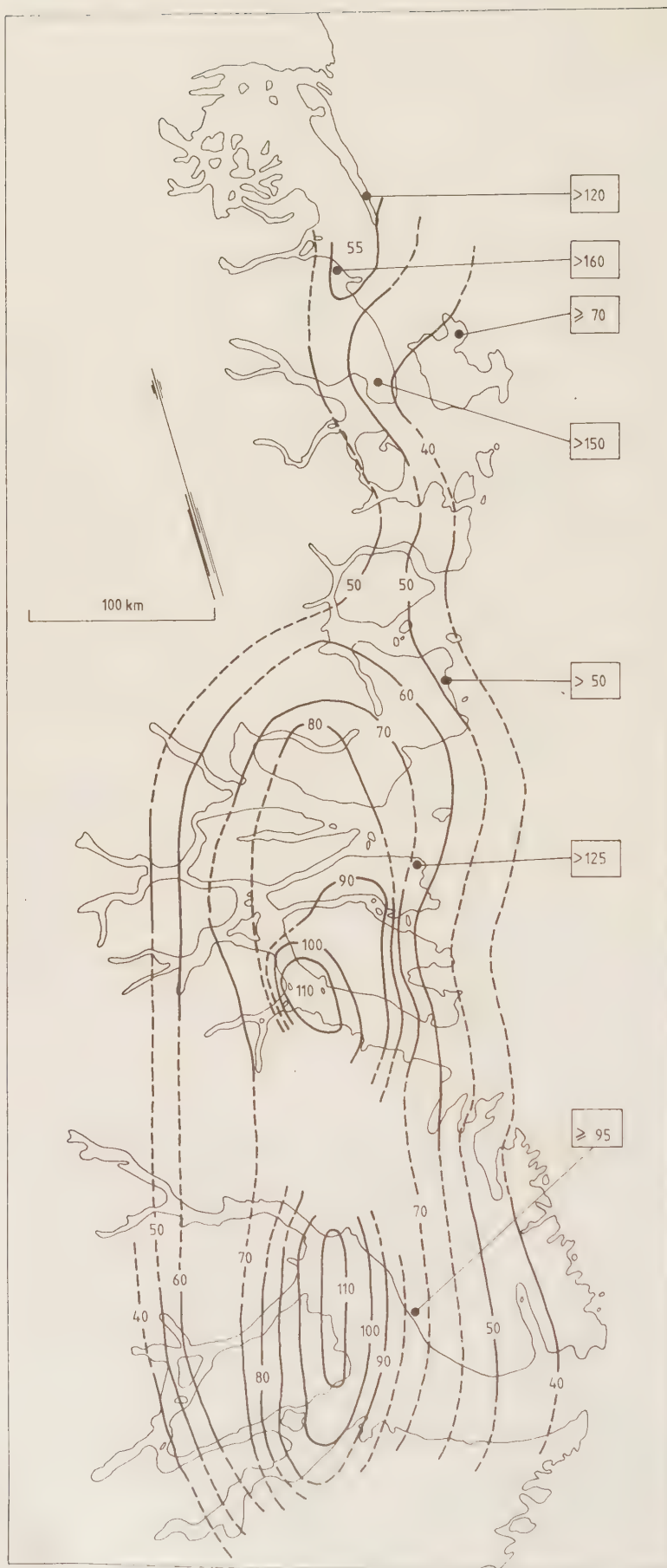


Fig. 3. Isobases (in m) for the marine limit c. 10,500 - 7000 B.P. Full lines show where the documentation is good, dashed lines indicate extrapolations. Scoresby Sund data from Funder (1978). Much information related to the area around Kejser Franz Josephs Fjord and Clavering Ø also from Funder (pers. comm.). In a few instances data from other sources have also been used. Marine levels of Middle Weichselian or older age are indicated by separate boxes. Altitudes given as > means silt with in situ shells, >> means traces of abrasion. For references see Fig. 10 in the original paper.

In Scoresby Sund deglaciation was under way already before 9800 B.P., whereas the oldest unequivocal deglaciation date from further north is as young as 9260 B.P. - from southern Hochstetter Forland. This indicates that the onset of deglaciation there could have started no earlier than 9500 B.P. It thus seems that deglaciation started somewhat earlier in Scoresby Sund than in the areas to the north. It should be noted, however, that the final upset of the glacial equilibrium in Scoresby Sund, which led to the rapid abandoning of the Milne Land moraine system, is dated to 9400 B.P., thus equivalent in time to the oldest datings from Fleming Fjord and Traill Ø and to the onset of deglaciation on southern Hochstetter Forland.

Thus the glaciation episode when the ice reached the B-line is a chronological equivalent to the Younger Dryas/Preboreal oscillations in Scandinavia. But in East Greenland the Late Weichselian glaciation maximum was reached then, so there is was more than only an oscillation during a longer period of general retreat from a maximum position further out, as was the case in Europe.

The date of the onset of the glacial advance towards the B-line is not yet known. But if it is linked to an increase in precipitation, following the penetration of Atlantic subpolar water into the Norwegian Sea region, then the beginning of this advance would seem to have taken place around 13 000 B.P. (However, recent data discussed in chapter 4 point to an even earlier advance).

Finally some comparisons with other areas are made, and it is found that there are similarities between the glaciation zone system in East Greenland and that in the eastern Canadian Arctic, notably on Baffin Island (where they are called weathering zones). But whereas from a morphological point of view the Late Weichselian B-line would seem to be an equivalent to the Zone III - II boundary on Baffin Island (Boyer & Pheasant 1974), the latter has been dated to about 100 000 B.P. However, the B-line may be complex, as in some places it consists of two moraines (and that problem is further dwelt upon in chapter 4).

The relationship between the East Greenland Late Weichselian and Early Flandrian glacier distribution and that of some other arctic areas is

also discussed, and it is noted that at least on eastern Baffin Island and Spitsbergen, and perhaps in West Greenland too, the glacial advances at that time were comparatively limited. As regards the timing of these advances it seems that within most of the eastern Canadian Arctic, and Spitsbergen, at least the culmination of glaciation and the onset of deglaciation was rather synchronous with that in East Greenland.

It is also noted that both in East Greenland and on Baffin Island and Svalbard there is one distinct younger marine limit, of Late Weichselian (Wisconsin) or Early Flandrian age, above which follows a series of more scattered and older deposits (Fig. 3). This is further treated in Paper 2 and briefly touched upon in chapter 4.

3.2 A glacial chronology for northern East Greenland

In this paper a glacial chronology for northern East Greenland is presented which probably includes the whole of the Weichselian and may even reach down into the Saale glaciation. The area covered by the study is roughly delimited by 75° and 76° N, stretching from Ardencape Fjord, Hochstetter Forland and Shannon northwards to Store Koldewey and the southern entrance to Dove Bugt. The field work was carried out during the Swedish-Danish Northeast Greenland Expedition 1976 by myself, Lena Adrielsson, Hans Bruch and Jan Mikaelsson.

The glacial chronology established for that area comprises three episodes of glacial advances, the Kap Mackenzie-, Muschelbjerg- and Nanok stadials, and three periods when more or less general deglaciation took place, the Hochstetter Forland- and Peters Bugt interstadials and the Flandrian Interglacial. This chronology is illustrated in Fig. 4 (= Fig. 3 in Paper 2), together with basic information on the extent of glaciation, marine limits and absolute dating.

During the Kap Mackenzie stadal glaciation was more or less total and the ice extended far out onto the continental shelf, as indicated by erratic boulders of western origin on all mountains near the coast, by deep glacial valleys cutting through Store Koldewey from west to east, and possibly also by some striae trending largely west - east on northernmost Hochstetter Forland, and large scale grooves on Store Koldewey.

This extensive and thick ice must have led to a very deep isostatic down-pressing of the land and thus during the following Hochstetter Forland interstadial, when a more or less general deglaciation seems to have taken place, the sea probably inundated larger areas than at any later time. Marine sediments from that period have been found up to altitudes of 120-160 m, and they have given ^{14}C ages in excess of 49 000 B.P. and Th/U ages between 70 000 - 115 000 B.P. Amino acid mean values for the ratio of isoleucine epimerization (Allo:Iso) were 0.25 for the free and 0.073 for the total (free + peptide bound) fractions.

After the Hochstetter Forland interstadial the ice advanced again, during the Muschelbjerg stadal, and covered some but not all of the older marine deposits with till. When, later, it melted it left behind a glacial

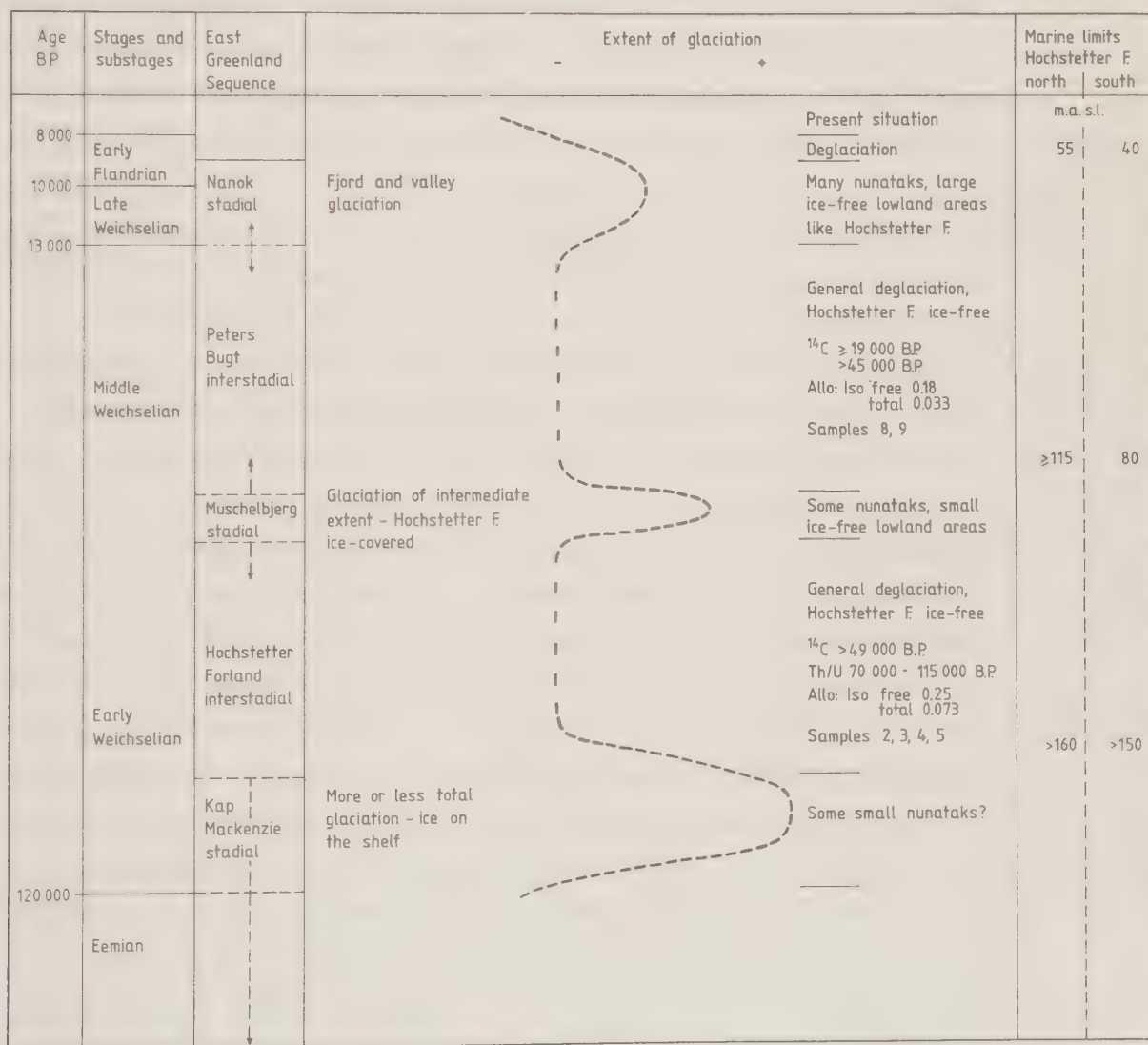


Fig. 4. Subdivision of the upper Quaternary in East Greenland between 75° and 76°N, with a glaciation curve and basic data concerning the extent of glaciation, the altitude of marine limits and the dating of interstadials. Amino acid results are mean values for each of the two interstadials.

landscape which, although morphologically more subdued than that dating from the younger Nanok stadial, is still largely intact. During the Muschelbjerg stadial glaciation was much less extensive than during the Kap Mackenzie stadial, and some nunataks and a few lowland areas remained ice-free.

As a consequence of the comparatively limited ice cover during the Muschelbjerg stadial the amount of isostatic downpressing must also have been smaller than during the Kap Mackenzie stadial. This seems to be confirmed by available sea-level data, that is by marine sediments and traces of marine abrasion from the following Peters Bugt interstadial. Also during that period a more or less general deglaciation seems to have taken place, and the sea reached altitudes of at least 115 m on northernmost Hochstetter Forland, about 80 m on southern Hochstetter Forland and probably 60-70 m on northeastern Shannon. Marine sediments from this period have been ^{14}C dated to more than 45 000 B.P. and the amino acid analyses (Allo:Iso) have given mean values of 0.18 for the free and 0.033 for the total fractions.

Later, during the Nanok stadial, the ice advanced again, but this time the extent of glaciation was even smaller than during the Muschelbjerg stadial and the glaciers were confined to the high mountains and the larger drainage basins. There were lots of nunataks within the eastern parts of the Caledonides and large ice-free areas existed east of there, including most parts of Hochstetter Forland and Shannon. But an ice-shelf covered the deep basin between Store Koldewey and Shannon, with a lobe extending southwards into the sound between Shannon and Hochstetter Forland. The upper and outer limit of glaciation within the drainage basins during this general period is the one identified throughout the fjord region as the B-line (Paper 1), and it constitutes a boundary between areas with fresh glacial landforms and largely unweathered glacially sculptured rocks, and areas with more subdued and weathered equivalents. In Peters Bugt and east of there on southernmost Hochstetter Forland the Ardencaple Fjord glacier deposited a distinct terminal and lateral moraine, partly on top of marine sediments from the preceding Peters Bugt interstadial. The onset of glacial advance during the Nanok stadial has not yet been dated, but in this paper it is suggested that it took place around or before 13 000 B.P. (and that problem is further discussed in chapter 4).

The onset of deglaciation after the Nanok stadial took place during the Early Flandrian, ^{14}C dated to about 9500 B.P. or slightly later. By 8800 B.P. the glacier lobe between Shannon and the mainland had disintegrated, by 8600 B.P. the valley glaciers on northern Hochstetter Forland had receded behind the present coastline, and by 8400 B.P. the front of the glacier in Ardencaple Fjord had withdrawn to west of Kildedalen. Extensive dead-ice topography in some of the larger valleys indicates that during the final stages of deglaciation the ice lost most of its dynamic properties.

The comparatively restricted glaciation during the Nanok stadial was reflected in a relatively small isostatic downpressing and nowhere within the study area does the Flandrian marine limit lie higher than 55 m. In contrast to the rather scattered traces of older marine inundation, Flandrian deltas, beach ridges and silt deposits are usually found all the way down from the distinct Flandrian marine limit to the present coast. The lowest Flandrian marine limit (35-40 m) is found on southernmost Hochstetter Forland and on northeastern Shannon. From there it rises northwards to about 55 m on northern Hochstetter Forland and at the entrance to Dove Bugt, and westwards to 50 m in Peters Bugt and eastern Ardencaple Fjord (cf. Fig. 3).

Regarding the different dating methods it is concluded that the ^{14}C ages of about 20 000 B.P. and older must be regarded as minimum ages, but that they at least show that the Peters Bugt interstadial began earlier than 45 000 B.P. and the Hochstetter Forland interstadial before 49 000 B.P. The Th/U datings also have to be considered with great cautiousness, but they seem to indicate an Early Weichselian age for the Hochstetter Forland interstadial, which would mean that the Kap Mackenzie stadial is also of Early Weichselian or even older age.

The pattern described here, with three glacial advances of successively smaller amplitude, is also known from other arctic areas, notably from Baffin Island, Ellesmere Island and Svalbard. Tentative and very preliminary correlations with the chronologies from Baffin Island (Miller et al. 1977, Andrews & Barry 1978, Andrews & Ives 1978) and from Svalbard (Boulton 1979, Troitsky et al. 1979) are suggested (also that problem is further discussed in chapter 4).

Finally it is noted that a strict correlation of glacial advances with the northward penetration of subpolar water into the Norwegian Sea region, that is to say mainly a correlation with increased precipitation, would seem to place all three glacial episodes discussed here within the Weichselian (Wisconsin), with glaciation maxima around 115 000 B.P., 70 000 B.P. and 10 000 B.P. (e.g. Ruddiman & McIntyre 1979).

3.3 Present and Middle Flandrian coastal morphology in Northeast Greenland

This paper is an attempt to assess the significance of two morphological zones with particularly distinct cliffs, beach ridges and shore bars, which occur along the coast of Northeast Greenland and date from Middle Flandrian and present times, respectively. The study is a by-product of glacial geological work carried out 1970-76.

Several authors have noted or assumed a recent submergence of various parts of Greenland, using archaeological evidence like marine inundation of Eskimo and Norse ruins, or tide-gauge observations and shore level displacement curves. Within the area of the present study marine erosion or inundation of Eskimo ruins has been noted by earlier authors as well as by myself. Earlier workers in the area also interpreted the occurrence of caverns and holes eroded into the rocks around the present high water mark on Ella Ő, and some very distinct beach ridges at the present shore on the same island, as a sign of sea level stagnation.

Field studies and studies of air photos confirm the above mentioned observations and show that within a broad zone from the outer coast and half way or more towards the Inland Ice the morphology of the present coastline gives strong evidence of a comparatively recent change in coastal processes. Further west the signs of this are less evident, but do occur. These indications are:

(1) The common occurrence of low cliffs at the present shore. - They are rarely more than 3-5 m high and usually formed in Quaternary deposits. Often a succession of beach ridges on a sloping plain is interrupted by a cliff at the present shore, indicating a break in the coastal processes where erosion comparatively recently got the upper hand. A special case are talus cones, which, due to marine erosion, have distinct cuts in their fronts.

(2) The occurrence of very pronounced beach ridges at the present shore. - On coastal plains with no or only comparatively insignificant beach ridges in the areas inside the present one, this is another indication of a recent change in the coastal processes, here resulting in a landward progradation of the accumulation of material.

Although these signs of coastal erosion are less pronounced in the west, which could be due to the stronger relief there with consequences for slope processes and deposition rates, they are of regional occurrence and thus cannot be explained by local factors. They need a regional explanation and there is then only one alternative to the submergence of land, or at least an isostatic-eustatic equilibrium, which could explain the onset of cliff erosion and the formation of the unusually large beach ridges. That alternative is increased wave erosion, most likely due to a diminished ice cover in summer, as proposed by Blake (1972, 1975) for arctic Canada. Data on the long-term changes of ice in the East Greenland Polar Current indeed indicate a decrease in its overall occurrence during this century, and as the climatic amelioration during the same time probably also led to slightly longer ice-free periods in the fjords, the exposure of the coasts could in fact have increased. But the same features occur along those shores of Spitsbergen where sea ice conditions are much less severe and wave erosion has probably always been more pronounced than in Northeast Greenland, and where consequently the hypothesis of increased exposure is difficult to apply, which speaks against a general application of this hypothesis. In combination with the tide-gauge measurements from West and Southeast Greenland, showing subsidence of land from the last century up to about 1950, this indicates that the morphology of the present shores in Northeast Greenland mirrors a transgression. Stratigraphic evidence for such a recent transgression seems to have been found on eastern Baffin Island (Dyke 1979, Andrews 1980), and probably these transgressions were induced by increased ice loads during the late Neoglacial (e.g. Kelly 1980 a).

In the typical case, on some gently sloping coastal forelands in the eastern part of the area, the distinct beach ridge or cliff at the present shore is followed inland by a sequence of less pronounced ridges. These may be well developed, but cannot compare with the modern one. Further inland this sequence of more normal ridges terminates at a cliff, or there is again one or a few very distinct beach ridges. Inside these another

sequence of more normal ridges may follow. There are also cliffs at higher altitudes, but they occur more erratically and their lateral extension is comparatively small. This zonation is illustrated in Fig. 5 (=Fig. 6 in Paper 3).

At the outer coast the foot of these distinct pre-recent cliffs may lie as low as 2 m above the present shore, and the upper limit of the zone where they occur at about 15 m. But the best developed cliffs and beach ridges are found between 6-8 m. The shore lines rise westwards and the one contemporaneous with the 6-8 m level on the outer coast lies at about 15 m in the innermost fjords where, however, the morphological zonation of the eastern areas has not been noted. This is probably an effect of the stronger relief and higher post-Weichselian isostatic recovery rate in the west.

The upper limit of this older zone with pronounced shore features has been dated to around 7500 B.P. and the most distinct cliffs and beach ridges within it, at about 6-8 m near the entrance to Kejser Franz Josefs Fjord, to about 6000 B.P. The upper limit was formed at a time when the very first signs of a slowdown in the rate of isostatic recovery of East Greenland began to show (Ten Brink 1974, Funder 1978), but when the eustatic rise of sea level was still continuing, and consequently that was the time when equilibrium situations between isostasy and eustasy became possible. It was also shortly after the break-up of the Laurentide Ice Sheet and at about the time when the first Tapes- and Littorina transgressions began in Scandinavia. At that time also the uppermost of some particularly distinct beach ridges and abrasion terraces on Svalbard were formed (Knape 1971), parallel with the beginning of an interlude with comparatively low emergence rate there (Salvigsen 1978).

The most distinct elevated shore lines in Northeast Greenland, those at 6-8 m from about 6000 B.P., correlate rather well, both in age and corresponding sea level, with the deposits laid down during the Vega Transgression, defined from a sequence in Vega Sund within the study area (5640 B.P.; Hjort 1973). They are also equivalent in time with the general period of maximum transgressions in Northwest Europe, and reasonably in accordance with dates from some of the best developed shore lines on Svalbard (6500 - 4800 B.P.; e.g. Blake 1970), with an influx of salt water

into a lake on Nordaustlandet (5550 B.P.; Hyvärinen 1969) and with a hump at about 6000 B.P. on Salvigsen's (1978) emergence curve. Nor do their age differ too much from that of some distinct shore lines on e.g. Ellesmere Island in arctic Canada (c. 5000 B.P.; Blake 1975) and it agrees with the time when sea level curves from southeastern Baffin Island sank below the present sea level (Dyke 1979). The timing with transgressions in northern Scandinavia, where Flandrian isostatic readjustment has been no smaller than in Northeast Greenland and increased exposure could not have played any role, seems particularly indicative of a transgressive background for these distinct coastal features in Northeast Greenland.

Thus local and circumstantial evidence indicate that the distinct morphology of the present shores in Northeast Greenland, and similar features in an elevated zone of Middle Flandrian age, are due to transgressions of the sea. The recent transgression was probably caused by Neoglacial isostatic downpressing, the older transgression (transgressions?) may have had an entirely eustatic background.

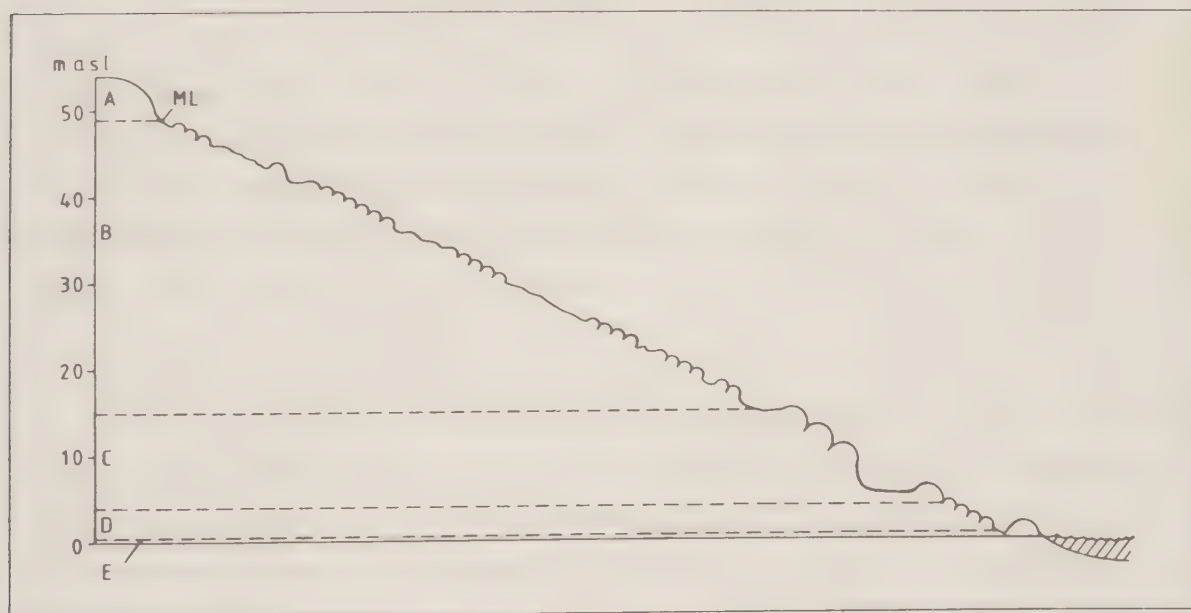


Fig. 5. An illustration of the morphological zonation on the outer coast of Northeast Greenland. (A) The area above the Late Weichselian - Early Flandrian marine limit (ML). (B) A mainly Early Flandrian zone with cliffs and beach ridges of moderate size. (C) The Middle Flandrian zone with its very distinct cliffs and beach ridges. (D) A mainly Late Flandrian zone with cliffs and beach ridges of moderate size. (E) The large beach ridge (or distinct cliff) at the present shore.



4. COMMENTS IN THE LIGHT OF SOME RECENT STUDIES

The three papers forming the basis of this thesis were written in 1978, 1980 and 1981, respectively. Recent publications have since thrown new light on the first two, which, along with the implications of two forthcoming papers on glacial geology and vegetation history in Northeast Greenland (Funder in press, Björck & Persson in press) is briefly discussed here.

4.1 The glaciation zones

Systems of weathering zones (= glaciation zones in the terminology of Paper 1) resembling those earlier described from Baffin Island and Labrador (e.g. Boyer & Pheasant 1974, Ives 1978) have now been found also on northeastern Ellesmere Island (England & Bradley 1978, England et al. 1981) and have been described again from Baffin Island (Dyke 1979). Many similarities exist between the different zonation systems in arctic Canada, and between these and the more primitive ones from Northeast Greenland and Norway, although the hitherto described morphological and weathering characteristics, and the absolute ages given to them in the

different areas are sometimes conflicting. However, as to the glaciation zone system of Northeast Greenland presented in Paper 1, it is evident that if the weathering criteria of Dyke (1979) were used there, it would indicate that rather large areas in the oldest glaciation zone, zone C, had been ice-free since the Early Weichselian or since pre-Weichselian times. This is what is indicated also by other data, e.g. from Hochstetter Forland and Shannon (Paper 2). But no areas totally without traces of former glaciation have yet been found in Northeast Greenland, although such are described from higher areas in different parts of arctic Canada. It should also be mentioned that Andrews & Ives (1978) stress the importance of a distinct break in the degree of weathering at the Baffinland Moraines, formerly known as the Cockburn moraines. The B-line in Northeast Greenland, i.e. the boundary between glaciation zones B and C, is roughly similar in age to those moraines and also characterized by a distinct break in the degree of weathering (Paper 1).

Weathering studies thus seem to be a promising working tool, even if papers by e.g. Sugden & Watts (1977), Blake (1978, Fig. 36.10) and Elven (1978) indicate that weathered surfaces, tors and felsenmeers could under certain circumstances survive a glacial inundation unscathed. But the abundance of glacial deposits, and of traces of glacial erosion, in the younger zones seems to indicate that glaciers which both settled silently on the ground and later just disappeared without ever getting dynamic enough to leave traces of some kind were not very common in these areas, or at any rate rarely or never became extensive.

4.2 The glacial chronology

In a forthcoming paper Svend Funder (Funder in press) presents evidence for the occurrence of marine deposits of Eemian interglacial age on Jameson Land in the Scoresby Sund district. As these deposits have not been overridden by ice it follows (1) that most parts of Jameson Land must have been ice-free since at least about 130 000 B.P., and (2) that the Kap Mackenzie stadial dates from the Saale glaciation. The latter was noted as a possibility already when that stadial was originally defined (Funder & Hjort 1973), and this idea was retained in Paper 2, but still only as a possibility. The logical consequence of a Saale age for the Kap Mackenzie stadial would seem to be an Eemian age for the Hochstetter Forland interstadial, and in fact the marine deposits on Store Koldewey which have

been tentatively assigned a Hochstetter Forland interstadial age contain shells of Cyrtodaria kurriana (sample 1 in Paper 2). This is a species occurring on the east coast today but which, except in this one place on Store Koldewey, has so far only been found as a fossil in the pre-Weichselian interglacial sediments near Kap København, at about 82°N in eastern Peary Land (Funder & Hjort 1980). There it is rather a type fossil.

A Saale age for the Kap Mackenzie stadial would also agree with the pre-Eemian (i.e. pre-substage 5e) age recently suggested for the Cape Christian interglacial sediments on Clyde Foreland, Baffin Island, and with the therefore probably increased age of the following Clyde Stade with its extensive glaciation (Fillon et al. 1981). Still it is not quite certain that this dating problem is solved that easily. No other supposedly interglacial molluscs have so far been found in the Hochstetter Forland interstadial sediments, and it should be kept in mind that the old erratics etc., which for want of possibilities to separate them chronologically have so far all been assigned to a Kap Mackenzie stadial, by all probability represent more than only this the latest phase of Pleistocene maximum glaciation on the different parts of the coast. This means that it is at least theoretically possible that the Kap Mackenzie stadial on Hochstetter Forland and Shannon, 500 km north of Scoresby Sund and 100 km closer to the Inland Ice than Jameson Land, could be younger than its present namesake in the south.

New evidence from Melville Bugt in Northwest Greenland (Kelly 1980 b) falls into the emergent picture, showing that also here the identifiable glacial advances have been of successively decreasing amplitude. An early ice sheet glaciation reached far into Baffin Bay, and was later followed by less extensive glaciation during the Kap Melville glacial episode, very tentatively dated to between 70 000 - 40 000 B.P. It could thus be a correlative to the Muschelbjerg stadial. The later glacial oscillations in Melville Bugt were probably of an even more limited extent. The similar picture of glacial events on northeastern Ellesmere Island, originally proposed by England & Bradley (1978) has recently been corroborated (England et al. 1981). It must be noted, however, that Blake (1977) has argued for an extensive late Wisconsin (Weichselian) glaciation on more southerly parts of eastern Ellesmere Island. This could be an effect of increased precipitation around the North Water polynya.

Since the publication of the new multiple glaciation chronologies from Svalbard (Boulton 1979, Troitsky et al. 1979), very similar to those from Northeast Greenland and parts of the Canadian arctic, the discussion of the Barents Sea ice sheet and related problems has again gathered impetus. New data collected during the Swedish Ymer-80 expedition have strengthened Boulton (1981) in his proposition of a complex glacial history also for Nordaustlandet, and in this he has been joined by Schytt and his colleagues (Schytt 1981 a, 1981 b), who identified 3 different sets of beach ridges on the northwest coast of Nordaustlandet, and it may be that only the lowest set is of Late Weichselian and Flandrian age. But Blake (1981) has argued against some of Boulton's (1979) conclusions, among other things referring to a shell sample collected at 82 m a.s.l. in outer Murchisonfjorden, dated to 11 180 B.P. According to Boulton (1981) and the working hypothesis of Schytt and his colleagues the marine limit of that age should be around 40-50 m a.s.l.

As regards the Late Weichselian glacial advance, new data presented in a paper by Björck & Persson (in press) indicate a complexity of the Nanok stadial on Hochstetter Forland. Their results are based on studies of two lake sediment cores from southern Hochstetter Forland and, among other things, they conclude (1) that the glacial advance to the main ridge of the Nanok stadial moraine in Peters Bugt must have taken place before 15 000 B.P., and (2) that after withdrawal of the ice front from this moraine followed a period of some thousand years when pollen from a mainly pioneer vegetation were deposited in a lake south of the ridge, and then (3) a very distinct deterioration of the climate took place about 10 000 B.P., leading to an interruption of the pollen production and to renewed glacial advance in Peters Bugt. According to the pollen data this later period ended about 9500 - 9200 B.P., which coincides well with the deglaciation from the Nanok stadial moraines there, as it has been dated in Paper 2. It is clear that in short parts the Nanok stadial moraine in Peters Bugt is double (as are several moraines along the B-line elsewhere in Northeast Greenland - Paper 1), and the lake from which one of the sediment cores was taken (with its oldest sediments ^{14}C dated to 12 960 B.P.) lies just between the main ridge to the north and a smaller ridge to the south. As the ice here came from the south it seems quite clear that the main advance during this stadial must have taken place well before 12 960 B.P., whereas the later advance, between just before 10 000 B.P.

and about 9500 B.P., may have formed the secondary ridge. That would mean that although the latter advance was of more or less the same amplitude as the first advance it must have been somewhat less dynamic, as pressure at the beaching points was smaller.

If the ^{14}C datings of these sediment cores are reliable, which seems to be indicated by the good agreement between ^{14}C dates and pollen stratigraphy, then the main phase of the Nanok stadial, according to Björck & Persson (in press) dating from before 15 000 B.P., approaches the classical European Weichsel maximum in time - although not in extension. Those sediments from the preceding Peters Bugt interstadial which have been overridden by the Nanok stadial ice, or where the contact to the sea was blocked by that ice, thus giving the maximum age of any glacial advance during the Nanok stadial, date from 35 400 B.P. or more and more than 45 000 B.P. (samples 6 and 9 in Paper 2). The sediments on southwestern Shannon which were dated to 19 000 B.P. lie outside the area covered by ice during the Nanok stadial (sample 8, which gave a technically finite age which, however, may still be a minimum one). So there is no tight time restriction backwards on the onset of the Nanok stadial. In this context it is very interesting to note that in his above mentioned paper Funder (in press) has redated the so-called Flakkerhuk stadial on Jameson Land to about 20 000 B.P., and that the advance towards the distinct Baffinland Moraines on eastern Baffin Island might also have begun about 20 000 B.P. These latter moraines are often complex, with an outer set dating from some time between 20 000 - 10 000 B.P. and an inner one from 9000 - 8000 B.P. (Andrews 1975, Andrews & Ives 1978).

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